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Heat flow and geothermal resources in northern Italy

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ABSTRACT

This paper gives an up-to-date overview of the surface heat flow and the geothermal resources of northern Italy on the basis of both already processed data and new pieces of information. Temperature data up to 7240 m depth, derived from exploration oil wells, were processed and thermal conductivity was estimated under any possible condition of depth. Radiogenic heat was evaluated by means of both natural gamma-ray logs and gamma-ray spectrometry measurements on core samples. These data together with information from previous investigations allow us to map the temperature at a depth of 2000 m in the Po Plain and to give a new picture of the surface heat flow throughout northern Italy. The most important hydrothermal systems occur in the Alps and the Northern Apennines, in areas where meteoric water leaks to shallow–medium depth and originates thermo-artesian springs. In the eastern sector of the Po Plain, important thermal anomalies appear to be controlled by the morphology of the deep carbonate formation. Thermal rather than forced convection can take place in this formation, which acts as a reservoir hosting low-medium enthalpy water.

Contents

1. Introduction

Studies on the geothermal resources in Italy have been traditionally focused on the high enthalpy fields scattered in the central-southern part of the country. Northern Italy, which includes the Po Plain, part of the Alps and the northern portion of the Apennines (Fig. 1), has instead drawn less attention, although it exhibits, in some areas, promising low-medium enthalpy resources. The main tectonic, structural and hydrothermal characters of northern Italy are well known from several geological and geophysical studies, both at the local and regional scale [1,2]. Among the regional studies, international scientific projects such as EGT [3] and ECORS-CROP [4,5] should be mentioned. Drillings carried out during several years of hydrocarbon exploration and regional seismic sections, integrated with magnetic and gravimetric modeling, have also given a contribution of paramount importance [6]. In the Mesozoic, the whole region experienced extensional events, which led to the formation of a wide carbonatic platform. Subsequently, the tectonic regime turned into compressive, producing, since the Oligocene, south-verging thrusts in the Southern Alps and north-verging thrusts in the Northern Apennines. At present, the carbonate platform crops out in the Southern Alps and is buried beneath the Po Plain.

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The pieces of available geological and geophysical information permit to distinguish the following tectonic units (Fig. 1):

- Southern Alps, related to the post-collisional (Oligo-Miocene) deformation of the Alps, consisting of thrust sheets, partly buried beneath the Po Plain Plio-Quaternary cover (SAB). The regional detachment surface of the thrust sheets corresponds to the top of the Eocene deposits. The few thermal springs occurring in this unit are often carbonic, of shallow–medium depth origin and with rather low temperature.
- Undeformed Po Foredeep (UPF), filled by terrigenous sediments deposited since Upper Eocene–Oligocene times and characterized by an impressive contribution of dismantling debris of the Alps and, only in recent times, of the Apennines. In the northern part of UPF, there occur outcrops of the carbonate formation together with Paleocene–Oligocene volcanic and subvolcanic products (Exposed Adriatic Foreland). With the exception of few thermal springs of minor importance, water, often salty, seems confined in the deep carbonate layers buried beneath the terrigenous sediments [2].
- Northern Apennines, consisting of thrusts sheets, which have developed since the Oligo-Miocene. Their external front is arcuate and buried beneath the Plio-Quaternary cover (NAB). The thermal springs occurring in this unit consist of either fossil, salty water, often associated with hydrocarbon reservoirs, or sulfate/sulfurous water related to the Messinian evaporitic series. Water can leak to shallow – medium depth only at few zones and originates thermo-artesian systems.
- Western Alps, including the deformed Alpine European foreland and the Pennine thrust sheets (ophiolitic unit cover and basement thrust sheets). Thermal springs generally originate from groundwater circulating in crystalline rocks. Important reservoirs occur also beneath the sedimentary cover of the Tertiary Piedmont Basin, structurally developed between the Western Alps and the Northern Apennines.

In this paper, we process deep temperatures from petroleum exploration wells of the Po Plain and carefully evaluate the rock thermal properties and the radiogenic heat in order to infer information on the surface heat flow. The new data, together with those so far available [2,7–12], allow us to describe the thermal regime and the characteristics of the main hydrothermal systems, and to outline an up-to-date picture of the geothermal resources of northern Italy.

2. Subsurface temperature

Temperatures recorded in hydrocarbon wells are the main source of information on the subsurface thermal state of the Po Plain. A huge number of drilling reports is contained in the open access file of the Italian Economic Development Ministry (Energy Department, General Direction for Energy and Mining Resources). Among the available well reports, we selected only those containing both temperature records and exhaustive lithostratigraphic descriptions. The position of the 98 selected wells is shown in Fig. 1.

As a whole, 295 temperature data from 280 – 7240 m depth range were analyzed. The dataset includes 277 temperatures measured at the hole bottom (BHT) and 18 recorded during drill stem tests (DST). BHTs are perturbed by the drilling mud circulating in the well. Thus, they had to be processed to infer the formation temperature. DSTs record the fluid (oil and gas) temperature, supposed to be in equilibrium with that of the surrounding rocks, and therefore no correction was necessary [13].

Depending on the information available in the well reports, we applied different methods to correct BHT data. When the well radius was unknown, we applied the empirical relation proposed by Pasquale et al. [10] based on the methods by Horner [14] and Cooper and Jones [15]. This approach was possible for most of the BHT data. When information on the well radius, shut-in time and mud circulation time was available, we applied the method by Zschocke [16]. In wells where repeated BHTs at a given depth were recorded together with the mud circulation time and the shut-in time, we used a method that envisages the well as a long hole of small diameter, drilled quickly and filled with water at temperature lower than the formation temperature (see for details [12]). In general, the correction for the circulating mud applied to BHTs ranges from 6 to 13 °C. The corrected BHTs and DST temperatures of the main tectonic units of the Po Plain as a function of depth are plotted in Fig. 2. The inferred thermal gradient varies little and it is minimum in NAB (21.7 mK m^{-1}) and maximum in SAB (24.5 mK m^{-1}).

Fig. 3 outlines the temperature distribution at a depth of 2000 m across the Po Plain. In the few wells that do not reach such a depth, temperatures were calculated by extrapolation. In the eastern part of the Po Plain (in correspondence of the UPF unit), the available data were too sparse to allow contouring (cfr Fig. 1). Thus, the temperature was estimated at the nodes of a regularly spaced grid (25 km × 25 km) from the thermal gradients of the sedimentary cover inferred from inversion of temperature data with a technique based on a lateral constant thermal gradient [17]. The stratigraphic column was schematized as indicated by

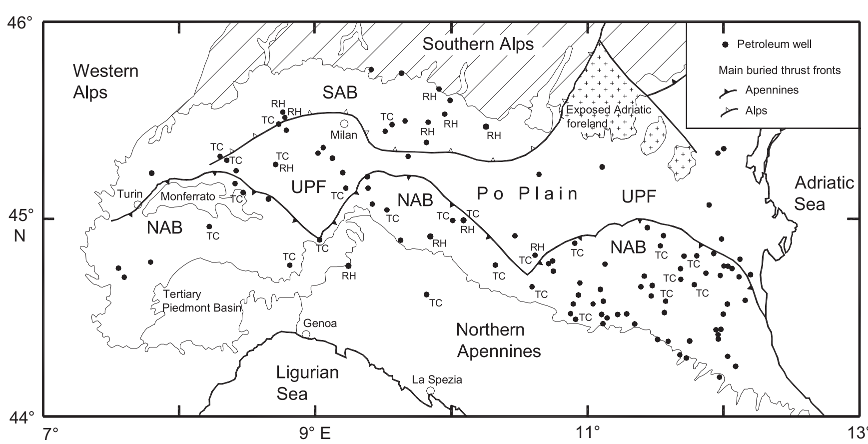

Fig. 1. Tectonic sketch of northern Italy and location of petroleum wells (full circles) providing temperatures, thermal parameters and lithostratigraphic information. Codes TC and RH next circles indicate wells which supplied cores for laboratory measurements and radiogenic heat data, respectively. NAB – Northern Apennines Buried, SAB – Southern Alps Buried, UPF – Undeformed Po Foredeep units.

Pasquale et al. [2] into four lithological units, by grouping formations having similar lithology and density: (a) an uppermost unit, characterized by marine sands, clayey sands and clays; (b) a sedimentary sequence essentially consisting of marls, silty marls and arenaceous marls; (c) a lithological unit mainly formed by argillaceous and marly limestones; and (d) the deepest unit corresponding to the carbonate rocks. The inferred thermal gradients are 20.2, 18.4, 52.7 and 13.5 mK m^{-1}, respectively.

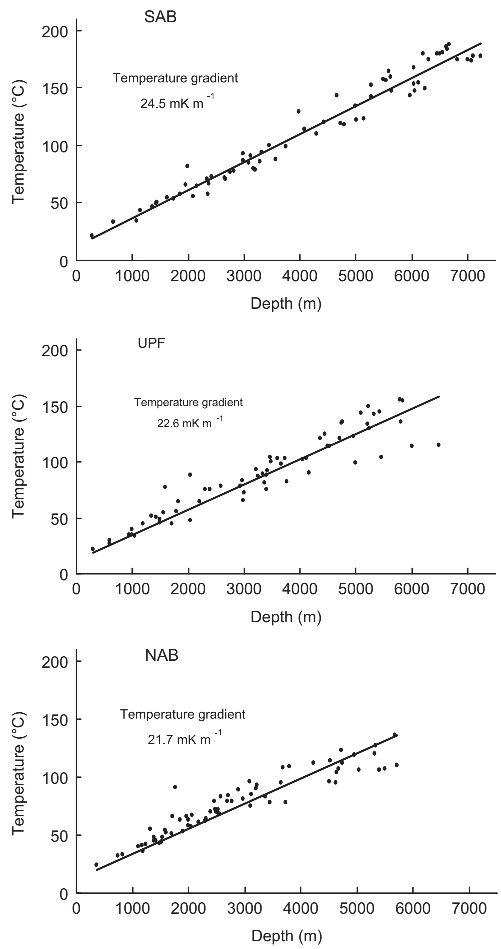

Fig. 2. Corrected temperatures versus depth in the different tectonic units of the Po Plain (see Fig. 1). The average thermal gradient is shown.

Most of the Po Plain is characterized at depth of 2000 m by a temperature ≤ 60 °C, with local minima (<50 °C) in correspondence of clastic permeable rocks (Fig. 3). In the central part, with the exception of a small area south of Milan, temperatures tend to be >60 °C. Values larger than 70 °C are observed in the eastern sector, in correspondence of the structural highs of the carbonate rocks, which are characterized by a reduced thickness of the Tertiary and Quaternary cover.

3. Surface heat flow

In 1980s, a few studies attempted to outline the surface heat flow in the investigated area [7–9]. These investigations were based on thermal datasets, mainly derived from petroleum exploration wells. Rigorous analyses to test the quality of these data, and to account for their meaning in relation to the different geological processes, such as sedimentation, thrusting and the deep groundwater flow, were however tackled only in recent times and led to the inference of surface heat flow in the western Po Plain [2,10–12]. In this study we analyze additional temperature data to extend the knowledge of the heat flow also in the eastern sector of the Po Plain. A preliminary analysis was carried out on the temperature data in order to circumvent those that might be affected by groundwater flow and thus unsuitable to heat flow determinations. In this regard, data from most of the oil wells drilled in NAB and UPF units in correspondence of the deep carbonate layer were rejected (see also Section 4.2). Thus we processed the temperature data from only nine wells that appeared dominated by a conductive thermal regime.

The heat flow was determined with the classical approach of the thermal resistance method. The thermal resistance R along the vertical between the surface and the depth d is

$$R = \Delta z \sum_{z=0}^{d} \left(\frac{1}{k_{in}} \right) \quad (1)$$

where k_{in} is the in situ vertical thermal conductivity at any depth interval Δz. The subsurface temperature in a horizontally layered, isotropic medium is related to the thermal resistance as

$$T_d = T_o + q_o R \quad (2)$$

where T_d is the temperature at depth $z=d$, T_o=13.0 °C is the ground surface temperature and q_o is the surface heat flow. T_o was assumed to decrease with elevation at a rate of 6 mK m^{-1}.

k_{in} was evaluated from lithostratigraphic data contained in the drilling reports by means of the technique proposed by

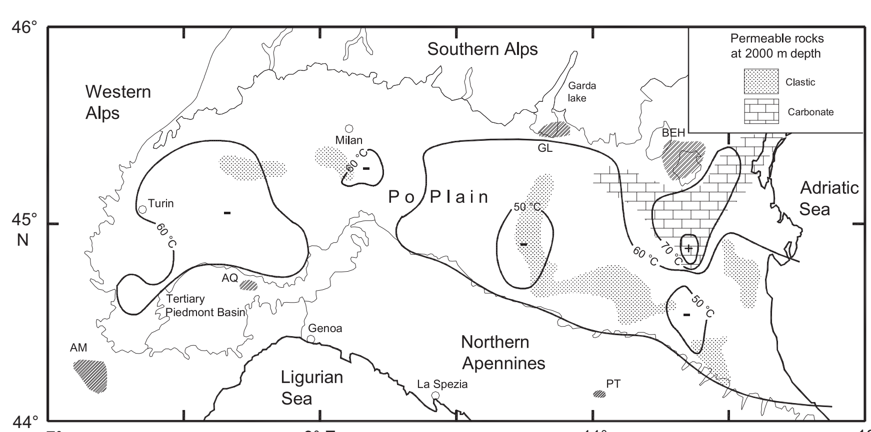

Fig. 3. Contour map of temperatures and permeable rocks in the Po Plain at 2000 m depth, as obtained from petroleum well data (see Fig. 1). Main hydrothermal systems: GL – Garda Lake; BEH – Berici Euganei Hills; AQ – Acqui Terme; AM – Argentera Massif; PT – Porretta Terme. Deep carbonate reservoir is also shown.

Pasquale et al. [11]. This technique is based on laboratory measurements of thermal conductivity and porosity, and mineralogical analyses (see also Fig. 1 for the position of the wells at which core samples were available), and assumes the geometric mixing model

$$k_{in} = k_m^{(1-\phi)} \, k_w^{\phi} \tag{3}$$

where ϕ is the porosity and k_w and k_m are the thermal conductivity of the water and the solid matrix, respectively. Porosity is assumed to decrease with depth z as

$$\phi = \phi_o \exp(-bz) \tag{4}$$

where b is the compaction factor and ϕ_o is the surface porosity. By expressing depth in kilometers, values adopted for ϕ_o and b were 0.180 and 0.396 km^{-1} in carbonate rocks, 0.298 and 0.461 km^{-1} in marls, silty marls and calcareous marls, 0.284 and 0.216 km^{-1} in sandstones and calcarenites, and 0.293 and 0.379 km^{-1} in shales, siltstones and silty shales, respectively.

Carbonate rocks, marls and sandstones were considered isotropic, whereas thermal anisotropy of the clay-rich lithologies (siltstones, shales and silty shales) was taken into account. In anisotropic rocks, the vertical thermal conductivity of the matrix, which decreases with depth due to the orientation of the clay and mica platelets during burial, was estimated by using the relation [11]

$$k_m = 2.899 - 0.251z \tag{5}$$

The water thermal conductivity k_w was assumed to change with temperature as [18]

$$k_w = 0.5648 + 1.878 \times 10^{-3}T - 7.231 \times 10^{-6}T^2, \quad T \le 137\ ^\circ\text{C} \tag{6}$$

$$k_w = 0.6020 + 1.309 \times 10^{-3}T - 5.140 \times 10^{-6}T^2, \quad T > 137\ ^\circ\text{C} \tag{7}$$

whereas the temperature dependence of the matrix conductivity was evaluated with the expression [19]

$$k_m = 1.8418 + (k_o - 1.8418)\left(\frac{1}{0.002732T + 0.7463} - 0.2485\right) \tag{8}$$

where k_o is the matrix conductivity at 20 °C. By taking into account the errors in correction for anisotropy, temperature and porosity, the estimated total uncertainty on thermal conductivity is 10 per cent [11].

Fig. 4 presents the geotherms computed by means of Eq. (2) and k_{in} calculated at the middle-point of 20 m intervals, after

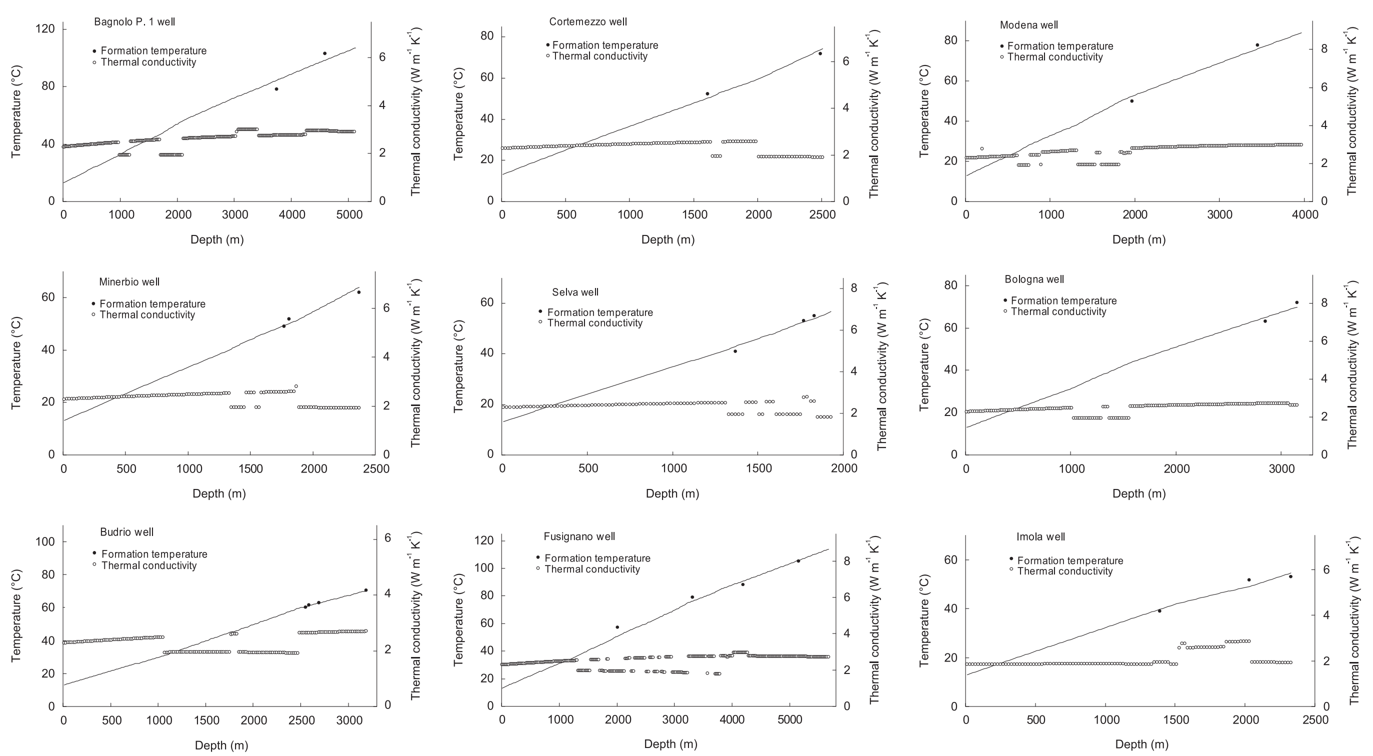

Fig. 4. Temperature profile and vertical thermal conductivity versus depth for the wells of Table 1.

Table 1
Average thermal conductivity k, radiogenic heat H and surface heat-flow values for the wells of the eastern sector of the Po Plain (see Fig. 6 for locations). The geographical coordinates, elevation and maximum depth of the temperature data are listed. HF observed heat flow; HF_r heat flow corrected for the radiogenic heat; HF_{rs} heat flow corrected for radiogenic heat and sedimentation.

Well code/name	Latitude (N)	Longitude (E)	Elev. (m)	Depth (m)	k ($\text{W m}^{-1}\ \text{K}^{-1}$)	H ($\mu\text{W m}^{-3}$)	HF (mW m^{-2})	HF_r (mWm^{-2})	HF_{rs} (mW m^{-2})
1HF Bagnolo P. 1	44.7790°	10.7091°	30	4650	2.60	1.02	47	52	65
2HF Cortemezzo	44.7578°	12.0257°	0	2485	2.32	1.06	56	59	69
3HF Modena	44.6159°	10.9366°	37	3456	2.60	1.10	46	50	60
4HF Minerbio	44.6111°	11.4715°	17	2357	2.30	1.05	49	51	63
5HF Selva	44.5838°	11.5725°	17	1828	2.32	0.99	52	54	61
6HF Bologna	44.5206°	11.2898°	42	3147	2.46	1.01	44	47	61
7HF Budrio	44.4893°	11.6909°	15	3180	2.27	1.14	40	44	55
8HF Fusignano	44.4436°	11.9807°	8	5144	2.46	0.97	43	48	68
9HF Imola	44.3842°	11.7480°	35	2333	2.05	1.23	36	39	51

incorporating the combined effects of mineral composition, anisotropy, temperature and porosity of the nine selected wells. Generally, in the uppermost kilometers, the compaction effect is larger than the temperature effect and, for the same lithotype, this causes an increase of conductivity with depth. Horizons of shale or silty shales are present at different depths and exhibit minima of conductivity. In these horizons, due to the presence of thermally anisotropic sheet silicates, conductivity is constant or decreases with depth. The observed surface heat-flow values, corresponding to the slope of the linear fit to data, together with the weighted average thermal conductivity in each well are listed in Table 1. The generally good fit between geotherms and observed temperatures confirms that groundwater flow is of negligible importance in these wells. However, Eq. (2) is based on the assumption that the radiogenic heat does not affect the temperature-depth distribution. This may introduce an uncertainty in the heat flow determinations. Moreover, the calculated heat-flow values ignore perturbations due to sedimentation and climatic changes. Therefore, corrections must be applied to the observed heat-flow data.

The radiogenic heat of the rocks was obtained by means of both natural gamma-ray logs, inferred with empirical relationships [12,20,21], and gamma-ray spectrometry measurements on core samples (Fig. 1). Fig. 5 shows an example of determination and Table 1 lists the average values of radiogenic heat deduced in each well.

Sedimentation in the Po Plain is of utmost importance. Its thermal effect could be in principle approached with different methods, e.g., by using a sudden deposition or a constant sedimentation model, which

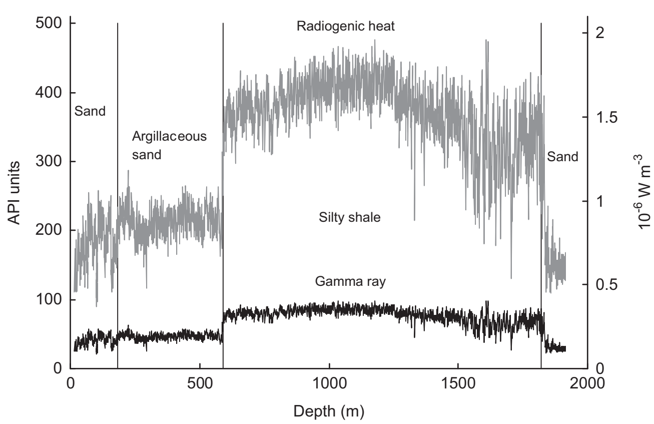

Fig. 5. Gamma-ray log and radiogenic heat of Carpaneto well (44.91° N, 9.84° E) (see Fig. 1).

of course may give different results [13]. Because several compressive tectonic phases, which involved shortening and overthrusting, have taken place, the sedimentation rate of the different deposition cycles, erosion or lacking of sedimentation and compaction are however difficult to quantify. Therefore, in order to evaluate the thermal effect of sedimentation, we used the simplified approach by Von Herzen and Uyeda [22], which is based on the assumption of a constant sedimentation rate. We modeled only the thermal effect of the most recent and important deposition cycle (Plio-Quaternary), considered as a single event, which took place on the Miocene formations acting as a basement. The model predicts a significant decrease of heat flow at sedimentation rates $> 10^{-4}$ m yr^{-1}. The correction to heat-flow data is on the average about 20 per cent, but at Fusignano well, located in the NAB unit where the cover thickness is about 5000 m, correction is as large as 29 per cent (Table 1).

We also evaluated the paleoclimate effect by means of the depth-dependent correction curve proposed by Majorowicz and Wybraniec [23] for south – southwestern Europe. The paleoclimatic effect as a response to five glacial cycles since 600 kyr ago with glacial–interglacial surface temperature amplitude of 7 °C was calculated for a model with homogeneous thermal conductivity (2.0 W m^{-1}K^{-1}), diffusivity (28.4 m^2 yr^{-1}) and basal heat flow (60 mW m^{-2}). Such past temperature changes predict a reduction in heat flow that smoothes with depth. The decrease in heat flow is as large as 5 mW m^{-2} at depth of about 1200 m and becomes negligible for depth larger than 2000 m. Since most of the temperature data available in the eastern sector of the Po Plain were recorded at depths larger than 2000 m (see Fig. 4 and Table 1), no correction for paleoclimate was then necessary.

The obtained surface heat-flow values, observed and corrected for the radiogenic heat and sedimentation effects, together with the average thermal conductivity and radiogenic heat are listed in Table 1. Fig. 6 shows the heat-flow map drawn by incorporating also the data recently revised by Pasquale et al. [12]. In zones where the spatial distribution of observations is adequate, we used the kriging technique to interpolate data. Where data were too sparse or even lacking (i.e., in the eastern part of the NAB and UPF units), the heat-flow pattern was estimated by taking into account the structural and tectonic affinity with the adjacent areas having a better data coverage.

In general, the heat flow of the Po Plain is 50 – 70 mW m^{-2} and increases towards the surrounding orogenic belts. Table 2 summarizes the average heat flow of the tectonic units in the investigated area. Heat flow varies across the Po Plain, being lower in the northern and southern units (66 ± 6 mW m^{-2} in SAB; 62 ± 6 mW m^{-2} in NAB) and larger (73 ± 4 mW m^{-2}) in the UPF

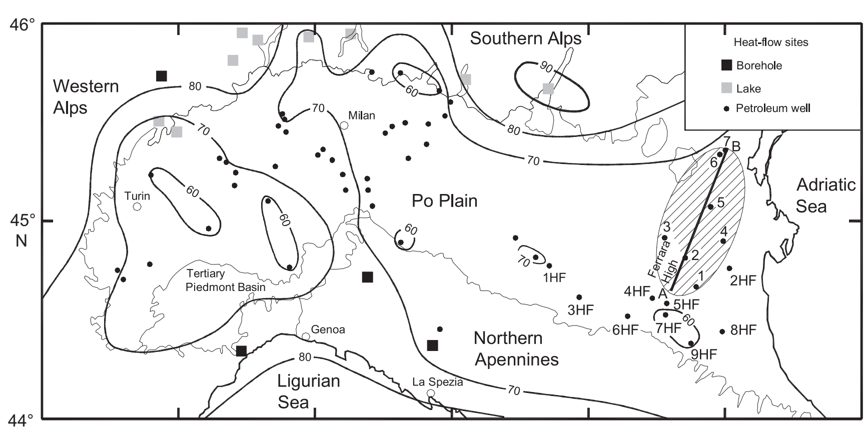

Fig. 6. Heat flow of northern Italy (isolines in mW m^{-2}). Available heat-flow sites used for contouring are shown. The new heat-flow values are labeled (see in Table 1). The hatched area includes the wells (1 – Consandolo, 2 – Ferrara, 3 – Cascina N., 4 – Cortevittoria, 5 – Villadose, 6 – Legnaro, and 7 – S. Angelo P.S.) penetrating the deep carbonate reservoir.

Table 2
Average surface heat flow and standard deviation for the tectonic units of Fig. 1.

Tectonic unit	No. of sites	Heat flow (mW m^{-2})
Western Alps	2	79 ± 4
Southern Alps	39	84 ± 12
Northern Apennines	3	70 ± 7
Po Plain	47	66 ± 7
– UPF	11	73 ± 4
– SAB	16	66 ± 6
– NAB	20	62 ± 6

unit. In the Northern Apennines, the heat flow is 70 ± 7 mW m^{-2}, whereas in the Alps it is larger and varies from 79 ± 4 mW m^{-2} (Western Alps) to 84 ± 12 mW m^{-2} (Southern Alps).

4. Heat in the groundwater flow

In the Po Plain, the deep carbonate layer lying at the base of the clastic sedimentary sequence plays an important hydrothermal role as it hosts hot water. In the mountain chains surrounding the plain, at several zones, meteoric water leaks to shallow–medium depth and originates thermal springs. Fig. 3 highlights the most important hydrothermal districts and the deep carbonate reservoir whose features are described in the following sections.

4.1. Advective hydrothermal systems

The Argentera Massif hydrothermal system (AM) presents the highest concentration of discharges in the Western Alps. The most important springs are Vinadio and Valdieri [24,25], which gush out within small areas (as a whole about 0.01 km^2) and show the highest flow rates and temperatures of the entire Alps chain. Besides, a few shallow wells (about 80 m deep) were also drilled at Vinadio to catch waters. The descending meteoric waters enter a granitic–migmatitic reservoir, where they equilibrate at temperatures of 95–130 °C, and discharge at a total rate of 0.007 $m^3\,s^{-1}$ with an average temperature of about 60 °C. The final upward flow is strictly related to sets of minor fractures, which are connected at depth to the major fault zones. For an average surface temperature of 5 °C, the total thermal yield of the AM springs is about 1.6 MW. According to mass balance calculations performed by Baietto et al. [26], the total discharge rate might be, instead as large as 0.05 $m^3\,s^{-1}$. In this case, the total thermal power should amount to about 12 MW. By testing with numerical modeling different thermo-hydraulic hypothesis, these authors also speculate that in the AM hydrothermal circuit thermal convection might coexists with advective flow.

The Sirmione hot springs are the main thermal manifestation of the Garda Lake (GL) hydrothermal system. According to Balderer et al. [27] water flows out at a NE–SE trending fault line affecting the carbonate bedrock. From the hydrogeological point of view, the Sirmione waters belong to a wide circuit developed within the carbonate formations cropping out between the Southern Alps and the Exposed Adriatic Foreland, and the recharge area is common to that of the Berici–Euganei Hills hydrothermal system. At present, the flow rate at the main spring, situated off the Garda Lake southern coast, is 0.004 $m^3\,s^{-1}$ and the temperature is 69 °C [28]. Other unexploited thermal springs occur at both sides of the main water discharge. Moreover, three wells, ranging in depth from 400 to 630 m and all reaching the carbonate bedrock, produce water for a total discharge of 0.039 $m^3\,s^{-1}$ and an average temperature of 64 °C. It is therefore possible to estimate that the total discharge at Sirmione has an order of magnitude of 10 MW.

The low temperature hydrothermal system of the Berici–Euganei Hills (BEH) extends over an area of about 23 km^2 in the northern part of UPF. In this area, several hot springs naturally flowed out in the past, but at present, due to the intensive exploitation, thermal water is by far extracted from more than 250 wells, most of which drilled for several hundred meters as far as the carbonate reservoir. The groundwater isotopic and chemical composition led to postulate a meteoric origin of the discharged water [29]. Water leaks in the Pre-Alps, at an elevation of about 1500 m, in correspondence of the carbonate formations of the Exposed Adriatic Foreland, and its deep circulation is controlled by regional faults. The water temperature at the discharge areas ranges from 60 to 86 °C and is very close to the rock–water equilibrium temperature. Therefore, if water warms up due to the thermal gradient of the UPF unit it is possible to estimate a maximum circulation depth of 2000–3000 m. The total flow rate in BEH is about 0.46 $m^3\,s^{-1}$ [30]. By assuming an average temperature of the water of about 70 °C and a surface temperature of 13 °C, we obtain for BEH a total thermal power extracted of 110 MW.

The Porretta hydrothermal system (PT) is located in the Northern Apennines. In an area of less than one square kilometer, there occur several springs whose water temperature ranges from 20 to 37 °C [31,32]. The geological setting is characterized by two main hydrogeological units, i.e., shales, forming the low-permeability unit, and a marly–arenaceous–silty turbidite with larger permeability, which acts as both reservoir and recharge of the hydrothermal system. In PT, the water discharge takes place through a deep regional tectonic fault that connects the reservoir to the surface. The fault permits the rise of overpressurized water accumulated at depth. No clear information is available about the rock–water equilibrium temperature. The total discharge for all the PT springs is 0.006 $m^3\,s^{-1}$ and the weighed average temperature is 32 °C. This leads to an estimated total discharge of only 0.5 MW.

Acqui Terme (AQ) is the only hydrothermal system where systematic well temperature logs and thermal conductivity measurements have been carried out and a remarkable number of thermal data is therefore available [8,33,34]. AQ is located in the Tertiary Piedmont Basin, in the suture zone between the Western Alps and the Northern Apennines. This basin consists of a thick Oligo-Miocene sedimentary cover of marls and embedded sandstone layers overlying a crystalline metamorphic basement. Over an area of a few square kilometers, numerous thermal springs occur with a total, natural flow rate of more than 0.015 $m^3\,s^{-1}$ and a thermal yield of 3 MW. The main hot spring has a temperature of 70 °C.

South of AQ, thermal data from the metamorphic basement of the Alps indicate that surface heat flow is 76 ± 5 mW m^{-2} [35]. This value can be considered as the background heat flow of the Tertiary Piedmont Basin. If an average bulk thermal conductivity of 2.2 W $m^{-1}\,K^{-1}$ is assumed for the cover and the basement, it follows that the regional thermal gradient is 34.5 mK m^{-1}. Geochemical data of the main hot spring argue for a rock–water equilibrium temperature of 108 ± 6 °C. Therefore, we can estimate that the maximum circulation depth of groundwater should be about 2800 m. On the other hand, temperature logs in wells show that an enhanced thermal gradient (>110 mK m^{-1}) appears to occur within a radius of 3–4 km around the thermal manifestations [33,34]. The enhanced gradient can be accounted for by water heating at depth because of the regional (background) heat flow and subsequently accumulating within a shallower reservoir.

The water flow path, the heating mechanism and the structure of the AQ hydrothermal system was quantitatively investigated by Pasquale et al. [36] by means of an analytical model of advective heat transfer within a fractured, water-saturated aquifer. Among the several trials performed under different hypotheses of aquifer inclination, thickness and hydraulic parameters, the model that better satisfies the available thermal constraints is that synthesized

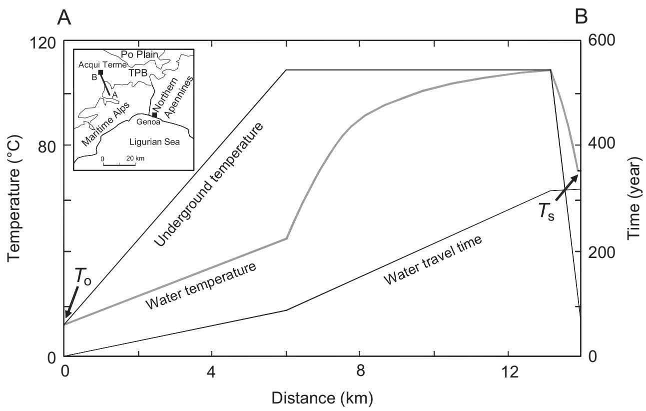

Fig. 7. Underground temperature, water temperature and water travel time of the Acqui Terme hydrothermal system along profile AB. $T_o = 13.0\ ^\circ$C and $T_s = 70.0\ ^\circ$C are the water temperature in the recharge and discharge zones, respectively. TPB – Tertiary Piedmont Basin.

in Fig. 7. The recharge of the hydrothermal circuit might be located 14 km south of the basin, where the crystalline basement of the Western (Maritime) Alps crops out [8,33]. From the fractured, recharge zone, meteoric water rapidly seeps at a moderate angle and warms up down to the maximum depth of 2800 m. Most of the heat is absorbed along the deepest, horizontal branch of the hydrothermal circuit. Since a large fraction of the system recharge might be widely dispersed and porosity should decrease with depth, a smaller effective thickness of flow and lower water velocity is hypothesized in the deepest branch. At the end of this branch, after a 13 km long path, the water temperature increases to about 108 °C (i.e., the maximum temperature of groundwater estimated from geochemical data).

The ascending section of the AQ hydrothermal circuit can be modeled with two branches. The first branch rises at an high angle from 2800 m to a reservoir located at intermediate depth. The upward flow of water might be relatively fast and occur through fractures with increased in porosity. The reservoir depth (about 850 m) and temperature (71 °C) can be inferred by extrapolating the thermal gradient observed in the wells next to the thermal area (70 mK m^{-1}). The final ascending branch of the flow path from the reservoir to the spring might occur through a relatively narrow, fracture zone (sub-vertical fault). The high flow rate of the spring suggests that the ascent should involve a negligible heat loss. Pasquale et al. [36] claim that a fracture modeled with a single sub-vertical cylindrical conduit with a radius of about 0.2 m in which water rapidly ascends into the surface with a velocity of 0.1 m s^{-1} can account for a temperature decrease by only about 1 °C during its final ascent.

4.2. Deep carbonate reservoir

In the eastern sector of the Po Plain, Pasquale et al. [2] found a considerable vertical variation in thermal gradient. The average value is about 21 mK m^{-1}, but in the layers above the carbonate reservoir the gradient is boosted to more than 50 mK m^{-1} (cfr Section 2). These authors demonstrate that such a vertical change cannot be ascribed to thermal conductivity contrasts, but a possible explanation could lie in the heat transported by water flow in the deep carbonate unit. Water flowing upwards can cause a decrease of thermal gradient in the carbonate reservoir and a significant increase in the overlying impermeable units.

The carbonate reservoir is hydraulically insulated and thermal convection can be the groundwater driving mechanism. The Rayleigh number analysis for average values of thermal gradient (13.5 mK m^{-1}) and thickness (about 5 km) of the reservoir indicates that a permeability larger than 3×10^{-15} m^2 is required for thermal convection to occur [2]. Such permeability is consistent with the minimum range of 5×10^{-17}–10^{-15} m^2 reported by Forster and Smith [37] and Manning and Ingebritsen [38]. In any case, if the bottom boundary temperature of the reservoir is much higher than that of the top boundary, convection can occur even at a Rayleigh number lower than the critical one [39]. This condition is described by the over-heat ratio, i.e., the ratio of temperature difference between the top and bottom boundaries and temperature of the bottom boundary. The over-heat ratio of the reservoir ranges from 0.32 to 0.61 and therefore is compatible with convection [40]. In the northernmost part of the eastern Po Plain, it cannot be excluded that water recharge from the carbonate outcrops in the Southern Alps produces convective/advective thermal regime.

Fig. 8 shows a cross-section with the patterns of the depth of the carbonate reservoir top boundary and its temperature together with the stratigraphic columns of the oil wells near the cross-section (see also Fig. 6). The sedimentary sequence mainly consists of terrigenous rocks, deposited on the Mesozoic carbonate successions cropping out in the Southern Alps. In NAB, both the sedimentary succession and the crystalline basement, underlying the reservoir, are involved in the Apennines thrusts faulting. The thrust front culminates in a structural high, known as Ferrara High, which is characterized by a reduced thickness of the Tertiary-Quaternary cover overlying the carbonate reservoir. The temperature distribution was calculated by assuming a surface mean annual temperature of 13.0 °C, and by extrapolating the thermal gradients of the four sedimentary lithological units overlying the crystalline basement (cfr Section 2). In general, the depth of the reservoir top and the calculated temperature have a similar trend, i.e., temperature increases where the carbonate top is deeper. The same occurs for the BHTs measured at the top of the reservoir. Notice that in two wells (Consandolo and Cascina N.) BHTs were recorded at depth greater than that of the top boundary, and, consequently, the temperature was extrapolated on the basis of the thermal gradient of the reservoir. However, the coherence between the patterns of the carbonate top and the measured temperature disappears between the Ferrara and Consandolo wells, which are located in the Ferrara High. In this zone, the measured temperature is much higher than the calculated temperature, thus indicating possible enhanced thermal convection.

5. Discussion

The subsurface temperature data for the Po Plain derive from records in deep oil drillings. Of the two kinds of available datasets, namely temperatures measured at the hole bottom and temperature recorded during drill stem tests, only the latter do not require further processing, because the fluid (oil and gas) temperature is regarded as that of the host rock (formation temperature). While drilling, the mud circulating in the well perturbs the formation temperature. This causes the deepest parts of the well to get cooled, while the ascending mud carries some of the heat from the deep formations and increases the temperature in the upper section of the well. Thus, with the exception of drill stem tests temperatures, data from oil wells were processed to infer the formation temperature.

We applied different techniques to properly correct bottom hole temperature data. In spite of the rigorous treatment, uncertainty of the inferred formation temperature may be still relatively large (say 3 °C). The same uncertainty should affect also the temperature distribution at 2000 m depth. However, the large depth at which temperatures were measured (up to 7240 m) decreases the uncertainty on the thermal gradients obtained, which can be instead considered of good precision [41].

Accurate processing of temperature data unaffected by groundwater flow gave new estimations of heat flow in the Po Plain. By

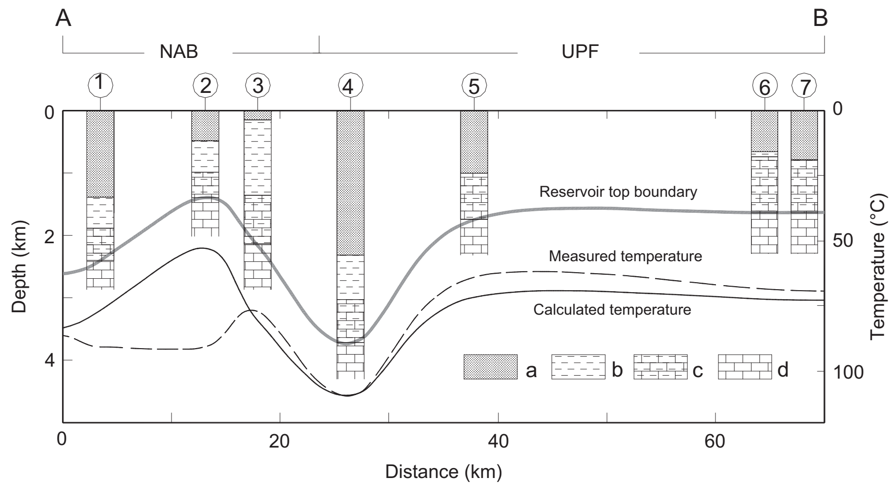

Fig. 8. Patterns of the depth of the carbonate reservoir top boundary and its temperature (measured and calculated) along cross-section AB. The numbers within circles indicate the well code number (see Fig. 6). Lithological units: a – marine sands, clayey sands and clays (Plio-Quaternary); b – marls, silty marls and arenaceous marls (Miocene); c – argillaceous and marly limestones (Paleogene); and d – mudstone, wackestone, packstone and dolostone (Mesozoic). NAB – Northern Apennines Buried unit, UPF – Undeformed Po Foredeep unit.

assuming that errors on formation temperature, surface temperature, well depth and thermal conductivity are 3 °C, 0.2 °C, 2.0 m and 0.25 W m^{-1}K^{-1}, respectively, for an average depth of 3500 m and a thermal conductivity of 2.5 W m^{-1}K^{-1}, the bias on each heat-flow determination is $\pm$10 per cent. The error on thermal conductivity is by far dominating the other contributions. Therefore, we addressed particular care to estimating thermal conductivity. The effect of thermal conductivity anisotropy, which is enhanced in sheet-silicate lithologies, was also taken into account. In these rocks, a geometric mean model was adopted [11] in which the thermal conductivity of the sheet silicates varies as function of the burial depth or compaction.

The new surface heat-flow values obtained are larger than the previous estimates [9,42]. The explanation for the enhanced heat flow lies in better estimates of thermal conductivity, the use of more accurate techniques to infer the formation temperatures and, secondarily, to the correction for the radiogenic heat, which was not taken into account in early studies. The new data, together with those obtained for the western sector of the Po Plain [12] and those available for the surrounding orogenic belts [43], allow redrawing the heat-flow pattern of entire northern Italy.

The surface heat-flow lateral variation seems to reflect the differences in tectonic history between the Po Plain and the surrounding orogenic belts. The heat flow varies from 64 mW m^{-2} in the Po Plain buried units to 73 mW m^{-2} in the Undeformed Po Foredeep unit. In the Alps the average heat flow is about 80 mW m^{-2} while in the Northern Apennines it is 70 mW m^{-2}.

In the light of the underground thermal state and the heat-flow pattern, heat advection is restricted to a few hydrothermal systems of the Alps and Northern Apennines, all characterized by low enthalpy water. Due to the generally low permeability of the rocks cropping out, in the Apennines water cannot penetrate to large depth and only a small number of thermal springs occur [44]. In the Alps deep flow systems can be instead encountered in all major geological and tectonic units. Fractured crystalline rocks (granite, gneiss), karstified formations, or porous sediments (sandstones, fluvio-alluvial deposits) may form aquifers.

The location and limits of the aquifers are highly dependent on local tectonic features [45]. Deep flow systems reaching 1–4 km depth mostly occur due to the presence of thrusts and subvertical faults having higher hydraulic conductivities. These fractures support the rise of deep water along the glacial and alluvial valleys, and complex mixing processes with shallow groundwater occur [46]. The emergence of waters can differ from one site to another; in some cases, only one large thermal spring occurs whereas in other cases, more than ten springs with different composition and discharge temperature can be present (e.g., Argentera Massif, Berici–Euganei Hills and Acqui Terme hydrothermal systems).

In the low Po Plain, at intermediate depth beneath the thick and impermeable sediments, a very slow groundwater flow may occur, due to mixing of water of shallow origin with deeper brackish water [47]. At larger depth, the carbonate formation plays a major hydrothermal role. There is evidence that in the Pedeapenninic zone it hosts a confined and salty aquifer, whose top can be as shallow as about one kilometer. Low-medium enthalpy water in these reservoirs is related to regional flow systems, characterized by cross-formational hydraulic continuity. The groundwater flow is caused not only by topographic gradient, but other phenomena, such as density differences in the water permeating rocks and thermal convection may act as additional driving forces [2].

Permeability inferred for the carbonate reservoir points to values well within the minimum range required for thermal convection to occur [2,37,38,48]. This hydraulic parameter depends on several factors, such as the aspect ratio of the flow domain (aquifer depth respect to width) and presence of faults or other discontinuities. Compared with the other sedimentary rocks, carbonates exhibit a wide variety of vertical and horizontal heterogeneities. In general, pores can be affected by mineral dissolution, replacement by other minerals and recrystallization, which take place during post-depositional processes. Porosity being equal, permeability can show a wide variation. Any way, primary porosity is often poor and secondary porosity (e.g., by fissure/fracture) of tectonic origin or related to unloading processes is more important. In the deep carbonate reservoir of the Po Plain, water flow might occur through micro- and macro-fractures, which might have been broadened by karst phenomena. At the Ferrara High, the larger temperatures observed (Figs. 3 and 8) are evidence of stronger thermal convection, which might be due to increased permeability caused by tectonism.

6. Conclusions

The subsurface temperatures and the surface heat flow obtained in this study form basic constraints to give an up-to-date picture of the thermal regime of the northern Italy. Geothermal resources mainly occur at several hydrothermal systems, scattered in Alps and Northern Apennines, and in the deep

carbonate reservoir of the eastern Po Plain. In the springs with lower temperatures and water flow rates (e.g., the Apennines hydrothermal systems), water flows at shallow depth and could be partially mixed with deeper hot water rising from the basement. Generally, the present-day knowledge of the hydrogeological and structural conditions only allows the formulation of conceptual model of the hydrothermal systems with the exception of the Acqui Terme system, where temperatures recorded in wells can be used as a tracer of the groundwater flow. In the eastern Po Plain, the deep temperature distribution indicates the presence of a confined, deep carbonate reservoir, occurring beneath the impermeable, low-thermal conductivity cover of clastic sediments, and highlights areas with thermal conditions potentially suitable for the settlement of heat extraction (district heating). The thermal and hydraulic properties of the sediment cover enhance the deep groundwater temperature. Such a deep reservoir is heated by the background regional heat flow and transports heat by convection.

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